SUBGLACIAL PRESERVATION OF VALLEY MORPHOLOGY AT AMUNDSENISEN, WESTERN DRONNING MAUD LAND, ANTARCTICA

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ABSTRACT

On the high altidude polar plateau of Amundsenisen, western Dronning Maud Land, East Antarctica, a subglacial valley, with a broad horizontal valley floor interpreted as a sediment floodplain or valley delta, was studied by radio echo sounding. In addition, a small, probably glacial, valley was mapped within the same subglacial massif. Basal ice temperatures were calculated using field data on precipitation, air temperature and ice sheet thickness. Discoveries of old landforms which have been preserved more or less intact beneath the former Fennoscandian and Laurentide ice sheets have received increasing attention during the last decade. The aim of this study is to investigate whether preservation of landforms occurs under the East Antarctic Ice Sheet, and to discuss under that climatological and glaciological circumstances preservation may take place. The results show that the ice sheet covering the investigated localities is frozen to bed, and therfore has an insignificant erosional capability. The observations suggest that a large-scale subglacial sediment deposit and a small valley formed by glacial erosion have survived beneath a cold-based ice sheet marginal zone for a long time period. The process of glacial preservation, recognized for bedrock features and tentatively observed for sediment accumulations, should act on similar large-scale landforms under any cold-based ice sheet, present or past. On the basis of existing studies of the age and stability of the East Antarctic Ice Sheet, a Middle Pliocene age is suggested for the preserved landforms. The presence of the presumed sediment-filled valley further indicates that no prolonged periods of basal melting have occurred at the Amundsenisen study area during the ice sheet history, which includes the Quaternary glaciation periods. Finally, calculations of basal temperature for localities at different altitudes within the same subglacial massif were used to demonstrate local altitudinal control of glacial preservation. © 1997 by John Wiley & Sons, Ltd.

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INTRODUCTION

A traditional opinion has been that glacial deposition and small-scale erosion landforms in previously glaciated areas are strictly the result of the final phase of the last glaciation. Such landforms predating the glaciation have often been believed to be erased by the subsequent ice coverage. This concept was, with few exceptions, considered a fundamental principle up until the last decade.

Glacial landforms are formed by erosion, transportation and deposition of eroded material. The most crucial factor regulating glacial erosion is basal sliding. Warm-based ice is known to slide over, and erode, its substrate (Kamb and La Chapelle, 1964; Sugden and John, 1976). Recent theoretical studies and field observations have indicated that cold-based ice may also slide (Shreve, 1984; Fowler, 1986; Echelmeyer and Wang, 1987), although the magnitude of sliding at subfreezing temperatures is extremely low. A few observations of cold-based ice not sliding over the bed have also been made (Goldthwait, 1960; Holdsworth and Bull, 1970). Cold-based ice can therefore be considered as essentially non-erosive (Robin, 1983; Drewry, 1986), with a possible exception in the formation of glacial striae (Shreve, 1984). The overall conclusion is thus that cold-based parts of ice sheets should have a very restricted influence on subglacial landforms. On the basis of glaciological theories, Schytt (1974) suggested that a major central part of the Weichselian Fennoscandian ice sheet may have

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been cold-based, a hypothesis later supported by Holmlund and Fastook (1995). A large proportion of the present-day East Antarctic Ice Sheet is also believed to be cold-based (Drewry, 1983).

An important way to obtain information about the ice sheet dynamics of Northern Hemisphere Quaternary glaciations has been to study the spatial distribution and orientation of glacial morphological features. In such studies, the assumption that a glaciation by necessity deforms and reshapes the landscape leads to a misinterpretation of the morphology and to conflicting results regarding ice sheet dynamics and configurations. The basically non-erosive characteristics of cold-based ice have to be taken into consideration.

The aim of this paper is to highlight the process and age relationships between an existing ice sheet and subglacial landforms, from a glaciological point of view. The paper addresses two main questions: (1) does preservation of landforms, including deposition forms, occur under the East Antarctic Ice Sheet and, if so, where does this happen on a local and regional scale: (this part includes a discussion of glaciological, topographical and climatological circumstances under which preservation may take place); and (2) what do the results tell us about the long-term thermal conditions of the ice sheet?

PREVIOUS WORK ON GLACIAL PRESERVATION

Studies of glacial preservation of landforms can be divided into two groups: (1) preservation documented in previously glaciated areas; and (2) observations of preservation beneath existing glaciers.

Preservation in previously glaciated areas

Recently, several papers have been published showing that small- to medium-scale depositional and erosional glacial landforms have been preserved beneath former ice sheets. For example, moraine features, drumlinizations, till ridges, boulder fields and fluvial patterns of a pre-Late Weichselian age have been reported from Scandinavia (Lagerbäck, 1988; Rhode, 1988; Kleman, 1992; Kleman and Borgström, 1994; Sollid and Sørbel, 1994) and from North America and Canada (England, 1986; Dyke and Morris, 1988; Boulton and Clark, 1990; Kleman *et al.*, 1994). Preserved deep weathering remnants have been reported from East Greenland (Funder, 1972), North America (Sugden, 1978) and Scotland (Hall and Sugden, 1987). A more detailed review of preserved pre-last stadial landforms in Scandinavia and North America is found in Kleman (1994). Several observations in these papers indicate that cold-based and warm-based basal conditions may vary on a local scale.

In the Dry Valleys region of Antarctica, Sugden *et al*. (1991) found unconsolidated sediments thought to have been overridden, but preserved, by a predominantly cold-based former ice sheet. Great antiquity has also been suggested for the origin of large-scale glacial erosion landforms in the Antarctic, found outside presently glaciated areas. Subaerial glacial cirques, possibly of a Tertiary age, have been described from Wright Valley in the Dry Valleys area (Selby and Wilson, 1971), while ice-free cirques and glacial valleys, with a suggested age younger than mid-Pliocene, have been reported from southern Prince Albert Mountains, Victoria Land (Verbers and van der Wateren, 1992; Verbers and Damm, 1994).

Glacial preservation by existing glaciers

Field evidence of landform preservation beneath present-day ice caps and ice sheets is rare. However, these type of observations are important because they can be linked to measurements or calculations of basal ice temperature, which controls erosion. Direct observations of small-scale landforms surviving cold-based ice coverage have been reported. For example, Goldthwait (1960) described patterned ground with lichens and mosses found beneath a Greenland glacier. Preserved vegetation, exposed by a retreating ice cap, was also reported from Baffin Island (Falconer, 1966) and Ellesmere Island, Canada (Bergsma *et al*., 1984) and from Kong Karls Land, Svalbard (Holmgren *et al*., 1984). Undisturbed Holocene beach ridges emerging from beneath cold-based ice caps have been observed on Storöya, Svalbard (Jonsson, 1983) and Alexandra Land, Franz Josef Land (Glazovskiy *et al*., 1992).

A few observations of preserved large-scale subglacial erosion landforms have been made using radio echo soundings of the Antarctic ice sheet. Cirques and glacial valleys, thought to originate from erosion by local wet-based mountain glaciers, were mapped on the inland side of the Trans Antarctic Mountains by Drewry (1972)

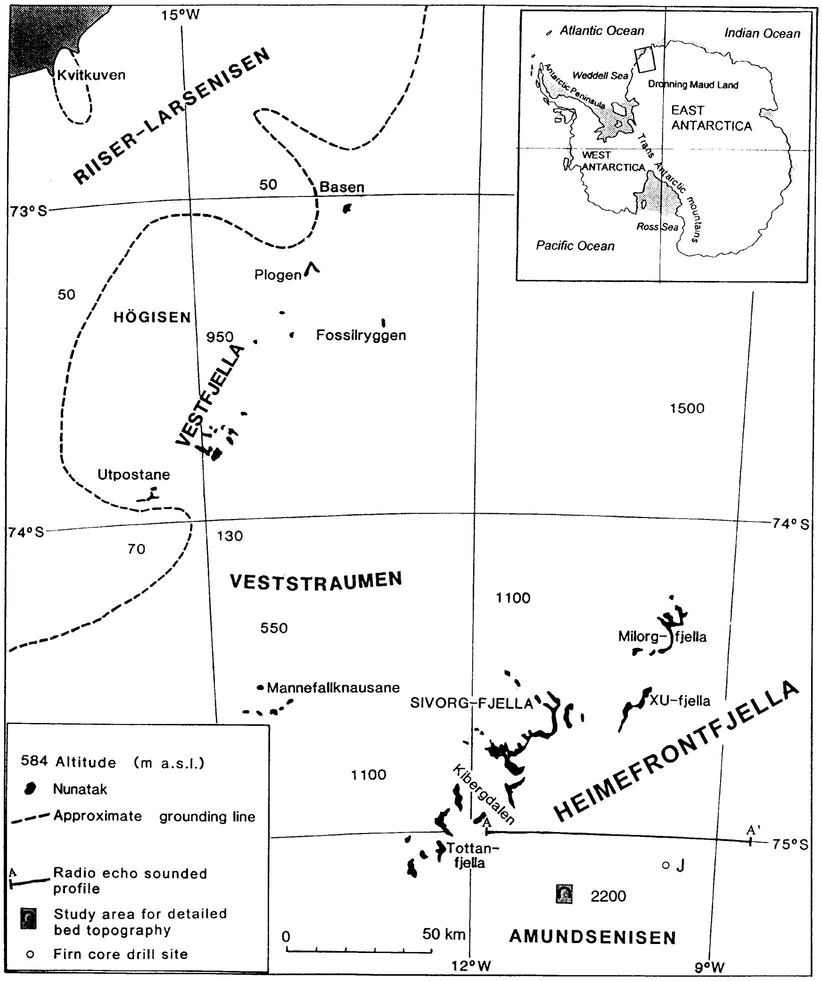

Figure 1. Location map. The 92 km long radio echo sounded profile along latitude 75°S is marked with a bold line (A–A′), while the location of the detailed bed topography study is denoted by a square. Climatological data are available from the firn core drill site marked with the letter J

and Calkin (1974), and close to nunataks in western Dronning Maud Land by Holmlund and Näslund (1994). On the basis of existing dating of the stability of the East Antarctic Ice Sheet, Holmlund and Näslund (1994) suggested that the age of the large-scale Dronning Maud Land glacial landscape was pre-Quaternary. In an extensive inventory study of alpine valley heads on the Antarctic peninsula, Haynes (1995) suggested that the main alpine glaciation in this area took place in the Miocene.

STUDY AREA

The study area is located in western Droning Maud Land, East Antarctica (Figure 1). Two parallel mountain ranges, Heimefrontfjella and Vestfjella, rise above the ice sheet surface within the region. The coastline, with the 100–150 km wide Riiser-Larsenisen ice shelf, is found northwest of Vestfjella. Between Vestfjella and Heimefrontfjella the ice sheet surface rises gently from 300 to 1100 m a.s.l. over a distance of 150 km. The Amundsenisen polar plateau, where the field work was conducted, is located southeast of Heimefrontfjella. The Heimefrontfjella range dams the ice flowing from Amundsenisen, causing a 1500 m drop in ice sheet elevation along a transect from southeast to northwest. Heimefrontfjella is made up of four major mountain blocks:

Figure 2. Photograph showing the glacial landscape in Sivorgfjella, Heimefrontfjella. The edge of the Amundsenisen polar plateau is seen to the right. The dominating alpine glacial relief on and around the nunataks is mainly due to erosion by wet-based local glaciers, predating the formation of the ice sheet (Holmlund and Näslund, 1994)

Milorgfjella, XU-fjella, Sivorgfjella and Tottanfjella. The largest valley cutting through the range, Kibergdalen, is located between Sivorgfjella and Tottanfjella.

The geological age of the Heimefrontfjella range is *c*.1100 Ma (Jakobs *et al*., 1993). The range is made up predominantly of Precambrian metavolcanic and metasedimentary rocks, in which granitoid plutons have intruded. In southeastern Sivorgfjella and Tottanfjella, where bedrock protrudes through the ice sheet closest to the investigation site, the rock is metamorphosed to an amphibolite facies. In some places within the range, Permo-Carboniferous sedimentary rocks and Jurassic lava flows overlie the metamorphic complex.

The subaerial bedrock morphology in Heimefrontfjella consists mainly of an alpine landscape, according to the landscape classification by Sugden (1978). The massifs are characterized by a local alpine relief, where the individual mountain block topography has constrained the flow of eroding cirque and valley glaciers. Common features are glacial cirques, U-shaped valleys and arêtes, highlighted in Figure 2. The glacial landscape on and close to the nunataks was most probably formed by local wet-based glaciers during a time when Antarctica had a climate warmer than at present (Holmlund and Näslund, 1994). This was tentatively suggested by Ahlmann (1944), who recognized the potential for having frozen parts of the ice sheet preserving relict morphology close to nunataks in Dronning Maud Land. Holmlund (1993) and Holmlund and Näslund (1994) provide a more detailed glaciological and geographical description of the Heimefrontfjella and Vestfjella area.

The two sites investigated are situated on the inland polar plateau of Amundsenisen (Figure 1). The first site consists of a 92 km long radio echo sounded profile along latitude 75°S (A–A′ in Figure 1), with the end points at 75°00′S, 11°40′W and 75°00′S, 8°29′W. The ice sheet altitude along the profile ranges from about 2000 m a.s.l. in the west to 2500 m a.s.l. in the east. This profile was sounded on two occasions during the field work. The second site covers an 18 km^2 area at 75°12′S, 10°55′W, also on Amundsenisen (Figure 1), which was radio echo sounded in detail for bed topography studies. At this site the ice sheet altitude is *c*. 2200 m a.s.l. The distance from the study area to the grounding line is 200 km and to the sea 330 km. Both sites are situated close to a local ice divide, caused by a subglacial mountain massif 40–70 km SSE of Sivorgfjella. The massif is distinguishable

as an undulating ice surface on the satellite image map Ritscherhochland SS 28–30 (Institut für Angewandte Geodäsie, 1988).

Temperature and accumulation measurements made in 1988/89 are available from a firn core drill site (at 75°04.78′S, 09°32.30′W), marked with the letter J in Figure 1. The firn temperature at a depth of 10 m was –28.3°C (Isaksson and Karlén, 1994). In areas without surface melting, the temperature at 10 m closely corresponds to the mean annual air temperature (Paterson, 1981). Accumulation measurements in the firn core showed a mean annual precipitation of 10·4 cm water equivalents at this site during the period 1955–1988 (Isaksson and Karlén, 1994).

RADIO ECHO SOUNDING METHODS

Two different radar systems were used for the acquisition of continuous ice depth and bedrock topography data. Along the latitude 75°S profile, data were collected during the austral summer 1991/92 with a range gated synthetic pulse radar developed by Environmental Surveillance Technology Programme, Norway (Hamran and Aarholt, 1993). It is a continuous-wave step-frequency radar using 201 frequencies evenly distributed over an adjustable bandwidth. The system has a high dynamic range, or sensitivity, and therefore a better potential for depth penetration than conventional impulse radars (Hamran *et al*., 1995). In the present study a pair of 155 MHz dipole antennae mounted on a Hägglund all-terrain carrier and a bandwidth of about 5 MHz were used. The transmitted output power was 10–20 W. The set-up has a theoretical maximum penetration depth of 3400 m. However, in practice the greatest possible ice depth recorded was 2300 m. The sampling interval was varied between 5 and 20 s, corresponding to a horizontal distance of 15 to 55 m between each recorded shot. The profile was positioned by Global Positioning System (GPS) measurements with a horizontal accuracy of ±100 m. All lat./long. positions in the paper refer to the WGS–84 datum. The absolute positioning in altitude, relevant when transforming values to altitude above sea level, had an accuracy of about ±200 m.

The ice depth profiles for the more detailed bed topography study, at the 18 km^2 area, were sounded in 1993/94. This was achieved using a high power output monopulse radar developed by the British Antarctic Survey. A pair of 60 MHz dipole antennae mounted on an all-terrain carrier module were used. The depth resolution was about 3 m. The return signal was, after amplification and filtering, recorded on a digital oscilloscope. Differential GPS measurements, with a relative accuracy better than ±5 m, were used for the positioning of the profiles. The accuracy of the absolute positioning in altitude was identical to the above survey. Both radar systems used laptop computers for operational control and data storage. In both cases, the antenna gain was strong enough to reduce the interpretation problem of the profiles into a two-dimensional problem along travel route. However, an exception to this in the latitude 75°S profile is discussed below.

RADIO ECHO RESULTS

The result from the radio echo soundings at Amundsenisen along the latitude 75°S profile is seen in Figure 3. The subglacial mountain massif southeast of Sivorgfjella is clearly visible in the left half of the figure. Several valleys are seen along the profile. For instance, the subglacial upper part of Kibergdalen is marked by an arrow 8–10 km from the starting point of the profile.

The valley discussed in this study is marked by an arrow in the central part of the figure, at 43–45 km from the profile starting point. It differs from other valleys along the profile by its horizontal valley floor and large valley depth. The subglacial mountains bordering the valley rise 1000–1400 m above the valley floor. The horizontal valley floor is located at an ice depth of 1850 m, corresponding to an altitude of 450 ± 200 m a.s.l. Based on the radio echo registration, the horizontal surface is not wider than 2000 m. Internal layers are seen within the upper half of the ice column, while a hyperbolic structure is found above the bed at the investigated valley, all providing information about present-day ice flow patterns and valley morphology (discussed further below).

Figure 4 presents the results of the bed topography investigation at the 18 km^2 site. The plot was constructed by interpolation of data between profiles of ice depth and ice sheet altitude. The valley is 3500 m long, 1500 m wide, and the main valley floor is located at an altitude of 1200 ± 200 m a.s.l. The steep bordering mountain ridges are 400–500 m high. The thickness of the ice sheet over the valley is 900–1000 m.

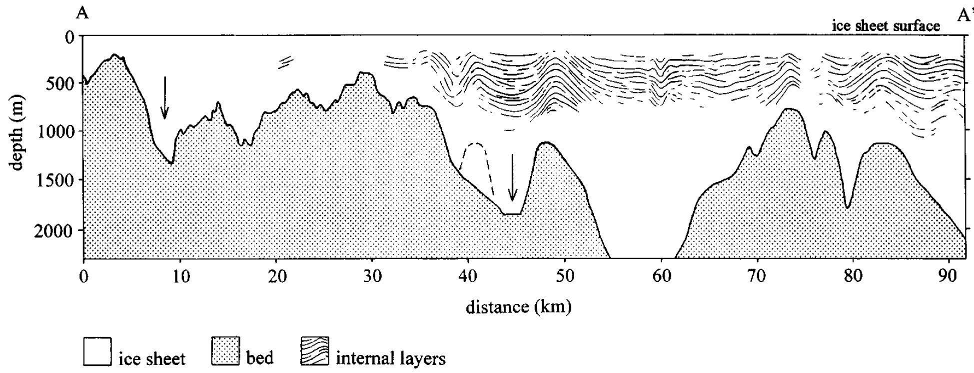

Figure 3. Profile of subglacial bed topography along latitude 75°S, with a vertical exaggeration of 10×. The bed topography has not been corrected for the ice surface topography. However, the surface relief of this polar plateau is low, making variations in the surface topography negligible with respect to the bed morphology. In the central part of the profile a deep, broad valley with a horizontal valley floor is found, denoted by an arrow. Today the valley is covered by 1850 m of ice. See the text and Figures 6 and 7 for an interpretation of the valley cross-section and its implications for the discussion on glacial preservation

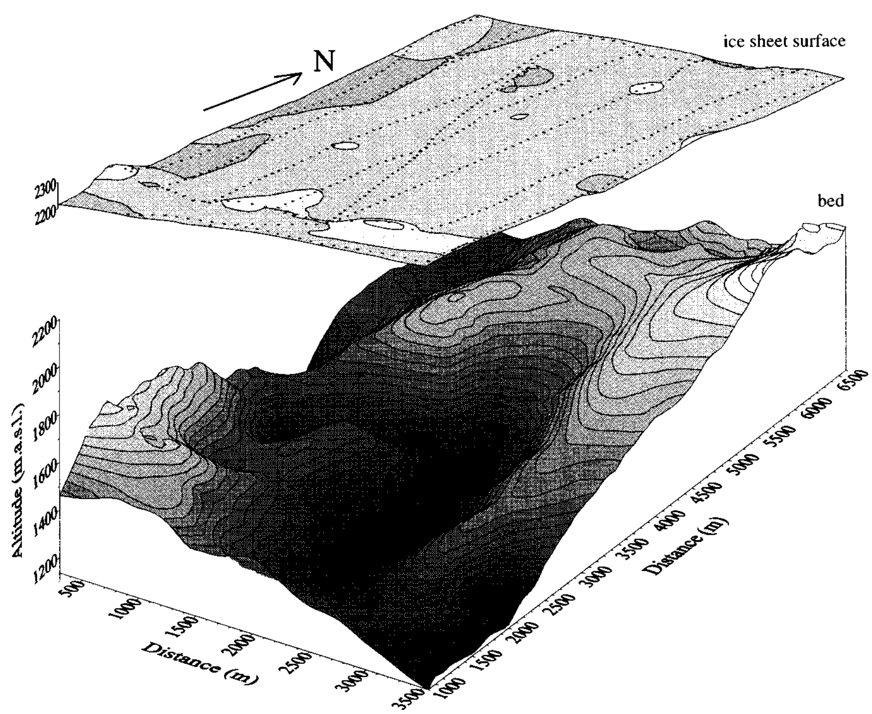

Figure 4. The subglacial valley system at Amundsenisen which was mapped in more detail, shown with a 2×vertical exaggeration and contour interval of 25 m. The bed surface has been corrected for the ice surface topography. Data on bed and ice sheet topography were collected with a spacing of 500 m or less along the profiles, shown by the dotted lines on the upper ice surface. The general valley morphology indicates that the valley was formed by erosion by a wet-based local mountain glacier. This site is used for a comparative calculation of basal temperature for discussing where, on a local scale, preservation of landforms may occur

TEMPERATURE DISTRIBUTION WITHIN THE ICE SHEET

As mentioned in the Introduction, glacial preservation may take place if the basal ice temperature is below the pressure melting point, preventing effective basal sliding and erosion. The vertical temperature distribution within the ice sheet has been calculated for the investigated sites by using the steady-state solution developed by Robin (1955), a solution commonly used in several studies (c.f. Siegert and Dowdeswell, 1995). In order to make this calculation, the following assumptions are made: (a) the ice sheet is in steady state (no change in ice thickness or vertical velocity, i.e. snow accumulation, with time); (b) heat from the internal deformation of the ice, produced mainly near the bed, can be treated as a flux in addition to the geothermal heat flux; (c) the firn layer is replaced by an equal thickness of ice; (d) the bed is horizontal; and (e) the basal ice temperature is below the pressure melting point. The last assumption means that the solution is valid for temperatures up to the pressure melting point. That is, if the ice is at a temperature lower than the pressure melting point, the formulas give a correct temperature, but in the case when the ice is at the pressure melting point the relationships break down. To study such a complex case, other formulas have to be applied in order to describe the thickness of the temperate basal layer (Funk *et al*., 1994). For the investigated sites on Amundsenisen, the Robin (1955) solution gave basal ice temperatures below the pressure melting point, and this solution was thus found to be applicable. When calculating the temperature profiles, the horizontal advection, or inflow of colder ice originating from higher altitudes, was set to zero. This type of solution was used because it is expected to apply in regions of low horizontal movement, such as close to an ice divide (Robin, 1955; Budd, 1969).

The temperature profiles in Figure 5 were calculated using the formula

$$\Theta_z=\Theta_s[\gamma_b\ 1/y(\mathrm{erf}\ y-\mathrm{erf}\zeta y)-(\alpha V\lambda/A)\ 2(Ey-E\zeta y)] \tag{1}$$

where Θ_z is the temperature at a specific elevation, z, above the bed, Θ_s is the temperature at the ice surface, γ_b is the geothermal heat gradient plus the internal frictional heat, α is the surface slope, V is the horizontal ice movement, λ is the vertical air temperature gradient, or lapse rate, and A is the vertical ice movement, i.e. accumulation rate (Budd, 1969). Θ_s and A were taken from the available firn core data. γ_b was calculated from

$$\gamma_b=\gamma_G+(\tau V/K) \tag{2}$$

where γ_G is the geothermal heat gradient, while the rest describes the heat produced by internal friction. A γ_G value of $0{\cdot}0228$°C m^{-1}, valid for Precambrian shields, was used (Budd, 1969). τ is the shear stress at the bed and K is the thermal conductivity of ice. A K value of $2{\cdot}1$ W m^{-1} K^{-1} was used (Paterson, 1981). The shear stress is calculated by

$$\tau=\rho gH(\sin\alpha) \tag{3}$$

where ρ is the density of the ice, g the gravity acceleration and H is the total ice thickness. The dimensionless parameter y in Equation 1 was calculated from

$$y=(AH/2\kappa)^{1/2} \tag{4}$$

where κ is the thermal diffusivity of ice. In the calculation a κ value of $37{\cdot}2$ m^2 a^{-1} was used (Paterson, 1981). The thermal diffusivity, thermal conductivity and density of the ice are assumed to be constant throughout the ice column (Funk *et al*., 1994). In Equation 1, the values of the error function erf x and Dawson's integral $E(x)$ were taken from tabulated values (Abramowitz and Stegun, 1970). ζ describes the fractional ice thickness, given by

$$\zeta=z/H \tag{5}$$

where z is the height above the bed. In Equation 1 a lapse rate of 1°C/100 m was used (Budd, 1969). For calculations of the pressure melting temperature of the ice at the bottom of the ice sheet, T_m, a linear relationship of the form

$$T_m=273-H/1503 \tag{6}$$

was applied (Remy and Minister, 1993). T_m is expressed in K.

Table I. General data and calculated pressure melting points and steady-state basal temperatures for the investigation sites on Amundsenisen

Site	Ice thickness, H(m)	Accumulation rate, A(cm a^{-1})	Annual air temp., Θ_s (°C)	Surface altitude (m a.s.l.)	Bed altitude (m a.s.l.)	Pressure melting point at bed, T_m (°C)	Calculated basal temperature (°C)
Sediment valley	1850	9·5	–28·3	2300±200	450±200	–1·2	–4·7±1·5
Glacial valley	1000	9·5	–28·3	2200±200	1200±200	–0·7	–12·5±1·5

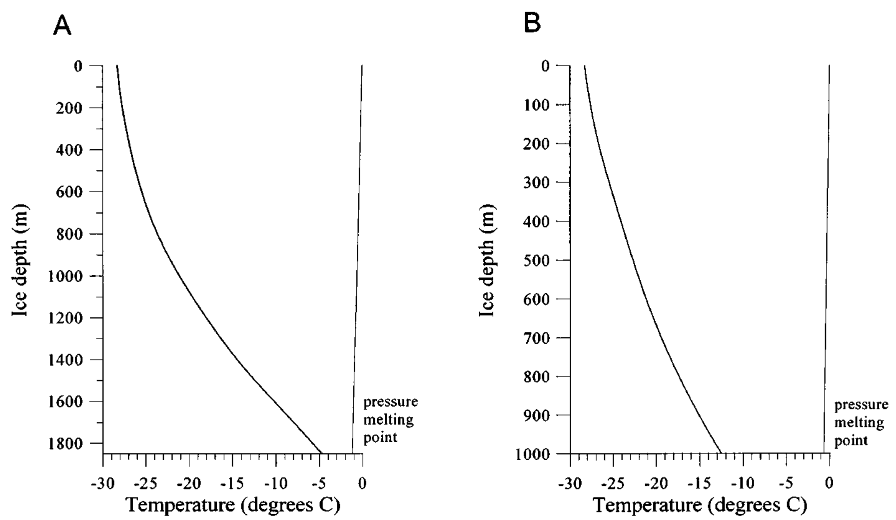

Figure 5. The vertical temperature distribution within the ice sheet at the locality of the horizontal valley floor (A) and at the site of the local valley in Figure 4(B). The calculated basal ice temperatures are about 3·5 and 12°C below the pressure melting point, respectively, preventing efficient glacial erosion

The climatic and physical parameters valid for the investigated sites, and the calculated ice temperatures, are summarized in Table I. The calculated vertical temperature profiles are shown in Figure 5. The investigated valley with the horizontal valley floor in Figure 3 has a calculated basal temperature of –4·7°C, whilst the valley in Figure 4 has a basal temperature of –12·5°C (Table I). The temperatures thus indicate cold-based ice at both localities. These calculated temperatures should be regarded as *indications* of the actual basal ice temperature. This is because the used boundary conditions at the surface and bed, i.e. the mean annual air temperature, accumulation rate and geothermal heat gradient, together with the method's inherent assumptions and the accuracy of the thermal parameters of ice, all affect the resulting temperatures. The accuracy of the calculated basal temperature values is therefore estimated to be ±1·5°C.

The ice thickness, air temperature and accumulation rate of snow are important factors for the englacial temperature distribution. During climatic changes, a thinner ice sheet, higher accumulation and lower air temperature result in colder basal ice, while a thicker ice sheet, lower accumulation rate and higher air temperature would give warmer basal ice. Since higher accumulation over a long period of time would also cause a thicker ice sheet, and vice versa, the interactions between these parameters are complex. Inclusion of the effects of a horizontal ice movement varying over time makes the above relations even more intricate. It is

beyond the scope of this paper to make a detailed analysis of these effects. Basal temperatures have therefore only been calculated for present steady-state conditions.

BED MORPHOLOGY, INTERNAL LAYERS AND ICE FLOW

Geomorphological interpretations of the bed surfaces in Figures 3 and 4 have been done in order to derive the most probable origin of the valleys. Because of the horizontal bottom surface of the investigated valley in Figure 3, the morphology is taken to indicate a valley containing sediments, with the horizontal surface constituting the surface of the sediment fill (Figures 6 and 7). Alternative interpretations for the horizontal

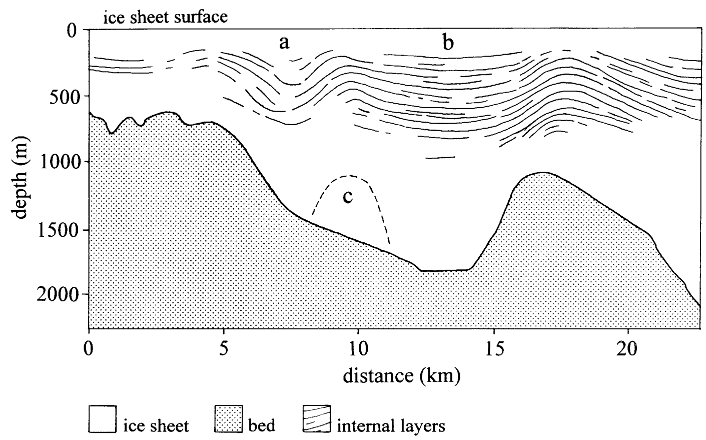

Figure 6. Partial enlargment of Figure 3, with a 5×vertical exaggeration. The subglacial valley morphology has been interpreted as a valley filled with sediments, constituting an old floodplain or valley delta. The ice sheet is here frozen to the bed, resulting in negligible glacial erosion. The observation suggests that the preservation of large subglacial sediment accumulations occurs beneath cold-based parts of the East Antarctic Ice Sheet. Accepting the floodplain/valley delta interpretation, the age of the sediment accumulation must be greater than the age of the ice sheet. See text for a discussion on the internal layers within the ice, marked (a) and (b), and the reflections at (c), and their relation to ice flow and bedrock morphology

surface could be a subglacial lake or a horizontal bedrock surface. These interpretations are less plausible though. For example, subglacial lakes beneath the East Antarctic Ice Sheet have been reported (Oswald and Robin, 1973; Robin *et al*., 1977; Drewry, 1982), but the calculation of basal temperature showed no indications of wet-based conditions at this site (Table I). In addition, a surface with free water at the base of the glacier would cause a significantly stronger reflection in the radio echo record than observed. A horizontal bedrock interpretation is highly unlikely because of the scarcity of such valley cross-sections elsewhere. Unfortunately the results from the radio echo soundings cannot be used to differentiate between drift and bedrock, and thus do not provide unequivocal proof for the sediment interpretation. However, juding from what normally causes such valley cross-sections, a sediment-filled valley still remains the most probable explanation for the mapped morphology.

The sediment fill could be either a till or of fluvial/glaciofluvial origin. Assuming a V-shaped valley cross-profile indicates that the sediment fill is about 150–200 m thick. Such thick till covers are rare, while finding a fluvial or glaciofluvial accumulation of this thickness in a valley in a mountainous region is not unusual. Furthermore, a horizontal upper surface covering an entire valley bottom is a highly unusual form for a till accumulation. The most common morphology comparable with the cross-section in Figure 6 is a valley with a floodplain, a valley delta or a valley sandur (in the United States called ‘valley train’). Floodplains are one of the most widespread fluvial depositional features to be found, existing in the valley of every major river and in most tributary river valleys worldwide (Ritter, 1986). Therefore, it is suggested that the subglacial landform is an old valley containing preserved sediments of a probable fluvial or glaciofluvial origin. Conclusions about

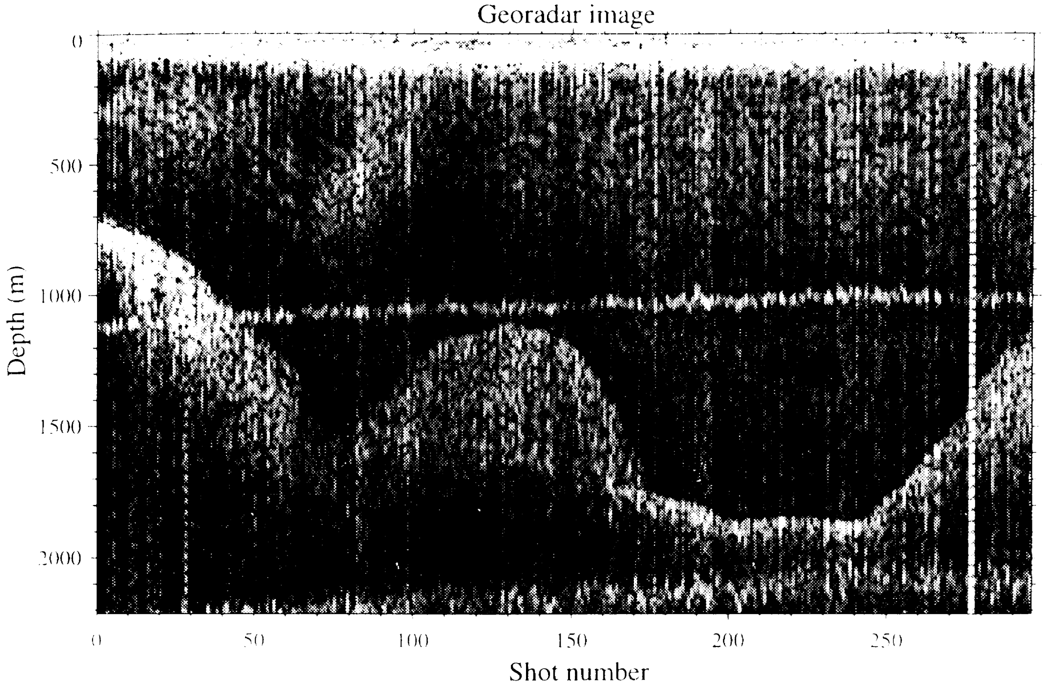

Figure 7. Example of a radio echo registration over inferred sediment-filled valley. The soundings were made with frequencies ranging between 155 and 160 MHz. The length of the x-axis corresponds to a distance of 12 km. The distance between each of the 296 shots is 40 m. Other registrations obtained over the same valley have a higher horizontal resolution with one shot each 2·2 m. The horizontal bed surface interpreted as sediment fill is here seen between shot 200 and 245. The nearly horizontal light bands at depths 1000–1100 and 2100–2200 m are disburbances caused by the radar equipment. For interpretation of the different features in the registration, see Figure 6 and the text

subglacial preservation of depositional landforms are undisputed whether the deposit is a till or glaciofluvial/fluvial fill, even though it would have implications for estimating the age of the form.

Radar soundings were made along the latitude 75°S profile on two occasions. This provided information for the valley morphology interpretation from two somewhat displaced, but parallel, profiles across the valley. The spacing between these is approximately 100–400 m. Data from both profiles are included in Figures 3 and 6. In the first of the profiles, the bed topography is shown by the solid line in Figure 6. In the other profile, a hyperbolic-like structure can be seen above the bed (Figure 6 at c, and Figure 7). This type of structure is common in radio echo registrations and originates from reflections from localized point targets (Harrison, 1970). The hyperbola is not a representation of the actual bed topography. However, it is likely that it results from reflections from the end of a steep mountain ridge located in the vicinity of the travelled profile (Figure 8). At that site the valley is probably split in two. Support for this interpretation comes from the structures of the internal layers within the ice sheet at an ice depth of 300–900 m (Figures 3, 6 and 7). Over the right-hand side of the valley in Figure 6 (at b), the internal layers are close to horizontal, while over the left-hand part (at a) they are strikingly concave. The concave pattern, having a vertical height of 250 m, most probably reflects higher snow accumulation and greater ice drainage at (a) than at (b). A greater drainage can produce a slight depression in the ice sheet surface, which leads to higher snow accumulation by wind redeposition, seen as a concave pattern on the radar registration. The association between local surface depressions and a relatively high snow accumulation appearing as concave patterns in snow radar profiles is described by Richardson *et al* (in preparation). The presumed difference in ice drainage between (a) and (b) supports the interpretation of two separate valleys at some distance from the travelled profile.

The difference in drainage between (a) and (b) further indicates that the separated valleys do not run in the same direction. The more active ice drainage at (a) indicates that this valley runs in a direction in concordance with the general ice flow direction in this area. The other valley seems to have an orientation less suitable for ice drainage. Although the flow rates in both valleys are probably very small owing to their location close to the ice divide, it is worth noting that the preserved sediments are located under the area which has relatively less flow of

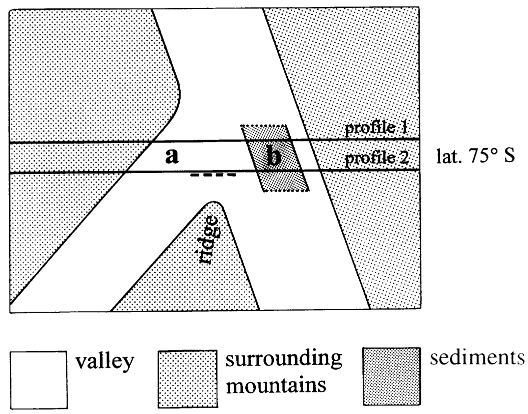

Figure 8. Schematic map of interpreted valley topology. A hyperbolic reflection from the end of the ridge was recorded along the dashed section in profile 2, due to the presumed close location to the ridge. (a) and (b) refer to Figure 6 and to the text. Because of low precision in the positioning of the two parallel profiles, it is not possible to determine on which side of them the dividing ridge is located. This means that the figure may be turned upside down. The sediments are preserved in the right-hand valley (as seen in the figure), interpreted as having relatively less ice flow.

ice (Figures 6 and 8). A higher flow increases the possibilities for having reached the pressure melting point at the bed due to heat produced by internal ice deformation.

The general features of the small valley in Figure 4 may be taken as incicating a glacial origin. The valley has one well-developed semi-circular backwall at the northern end of the valley. A less distinct semi-circular backwall can be seen forming one side of a sharp arête towards another valley in the west. The valley floor displays several overdeepenings of which the lowermost is the most pronounced. Furthermore, the steep valley sides and the size of the valley are morphologically similar to small local glacial valleys.

GLACIAL PRESERVATION AND ICE SHEET DYNAMICS

The basal temperature at the site of the sediment-filled valley is about 3·5°C below the pressure melting point (Table I, Figure 5A). The present conditions therefore do not allow significant basal erosion but rather favour glacial preservation. The presence of the sediments, and the results from the calculation of basal temperature, suggests that glacial preservation of delicate depositional landforms does take place beneath cold-based parts of the East Antarctic Ice Sheet. Accepting the fluvial/glaciofluvial origin of the sediments, the results further indicate that no extensive periods of basal melting and erosion have occurred at this site during the Amundsenisen Ice Sheet history. However, considering that the present basal ice temperature is only 3·5°C lower than the pressure melting point, the basal ice at this locality may have been warm-based for short periods during the evolution of the ice sheet. During such periods, restricted erosion could have taken place, which may have removed some of the sediments of an originally thicker sediment pack. The planar morphology of the landform may have been basically unaffected provided that the erosion rate was sufficiently low, for example due to a low pore-water pressure. However, the ice flow during warm-based periods must have been strongly governed by the valley geometry, giving the highest sliding velocity and erosion rate in the centre of the valley cross-section. This is in contrast to, for example, the more sheet-like flow in broad ice streams. Extensive periods with warm-based conditions at the sediment-filled valley would therefore most likely have transformed the horizontal bed topography into a more U-shaped valley cross-profile. It is thus likely that the ice coverage has remained cold-based and non-erosive at this site during most of the Quaternary (see also below).

The inferred preserved sediment fill is probably not the only conserved old feature within the studied subgacial mountain massif. The calculated basal ice temperature at the bottom of the glacial valley in Figure 4 is 12°C lower than the pressure melting point (Table I, Figure 5B). This shows that cold-based non-erosive ice covers the glacial valley at present, and thus that the valley has not been formed by the present-day ice sheet. Most likely it originates from erosion by a small wet-based mountain glacier, existing at a time when the Antarctic climate was warmer than today. This is in line with the results of a thorough study of subglacial cirques and U-shaped valleys around Vestfjella and Heimefrontfjella by Holmlund and Näslund (1994).

At both investigation sites on Amundsenisen, the cold-based conditions of today are to a large extent due to the location under an ice divide at high altitude, providing low ice movement and low annual air temperatures. The conditions for having cold-based ice in this area were probably similar or more favourable during the Quaternary glaciations. On those occasions, the distance to the sea was greater than today, firstly because sea-level dropped, forcing the grounding-line further out on the continental shelf, and secondly because of a more extensive sea ice coverage. Thus, the increased distance to the moisture source for precipitation resulted in reduced snow accumulation over the ice sheet interior. The lower air temperature also contributed to the moisture starvation. The ice sheet in the studied area may therefore have been somewhat thinner during glacials (although out at the former grounding line the ice sheet must have been significantly thicker than today). A numerical Antarctic ice sheet modelling experiment by Fastook (1995) indicated a lowering of the ice surface in the Amundsenisen area of 0–100 m during a Last Glacial Maximum scenario. The lower air temperature, together with a less thick ice sheet interior, probably acted in favour of lowering the basal temperatures during the Quaternary glacials. Therefore, cold-based conditions most probably prevailed at this locality during the Quaternary, preserving the mapped subglacial landforms.

All non-glacial landforms beneath the glacier must have formed before the ice sheet developed. Therefore, accepting the floodplain or delta interpretation, the sediment accumulation must predate the ice sheet. This is also the case for the mapped glacial valley formed by a small wet-based mountain glacier. Several studies imply that the East Antarctic Ice Sheet configuration has been stable since the Middle Miocene, i.e. over the last 11–15 Ma (Kennett, 1982; Marchant *et al*., 1993a,b), while other studies suggest a major deglaciation of East Antarctica during the Middle Pleiocene, 3 Ma ago (Barrett *et al*., 1992). From these studies, a minimum Middle Pliocene age is suggested for the preserved landforms.

A demonstration of local altitudinal control of glacial preservation may be done by comparing the two Amundsenisen investigation sites. The glacial valley in Figure 4 is situated within the same massif as the sediment-filled valley in Figure 6, but at an 800–900 m higher altitude. This has resulted, owing to the lesser ice thickness, in an 8°C lower basal ice temperature at the glacial valley (assuming the same accumulation rate over both sites) (Table I). From this example we see that it is probable that morphology at higher elevations has been less affected by ice sheet erosion compared to sites at lower altitudes. The conclusion is that preserved, relict landforms, such as the small glacial valley, are more likely to be found at higher altitudes within the subglacial massif. In contrast to these cold-based high altitude sites, areas where the basal ice is at the pressure melting point can probably be found in the deep trenches between massifs. Within these intra-massif ice drainage channels, glacial erosion may be an active process today because of the large ice thickness and relatively large ice movement. The local ice thickness and ice velocity are therefore important factors in governing the basal temperatures, in turn determining where ice may preserve or erode its substrate.

Although located under a local ice divide at the high inland polar plateau, the sediment-filled valley at Amundsenisen is situated 200 km from the present grounding line, which is within the marginal zone when considering the ice sheet as a whole. This observation supports the idea that preservation of landforms may occur not only 'under ice divide location' (Boulton and Clark, 1990), but also because of cold-based conditions under more peripheral parts of an ice sheet (Kleman, 1992). To some extent this demonstrates the difficulty of using broad generalizations about ice sheets when discussing the occurrence of regionally preserved landforms.

The preserved sediment fill is a depositional landform that suggests that glacial preservation of fragile features occurs beneath present cold-based ice sheets. Although it is not an example of a glacial directional landform that could be used for reconstructions of deglaciation patterns in previously glaciated areas, it is a relatively large-scale landform. Its size may be compared with large drumlinoids or flutings, directional landforms that are used for interpretations of previous glaciations in the Northern Hemisphere. The

preservation process, in the study tentatively observed for sediment landforms, therefore should operate on large sediment landforms beneath any cold-based ice sheet, regardless of hemisphere or point in time.

CONCLUSIONS

Field observations suggest that cold-based parts of the Amundsenisen Ice Sheet, East Antarctica, have the capability of preserving delicate old sediment deposits as well as large-scale bedrock features. Calculated basal temperatures indicate cold-based conditions, and thus negligible glacial erosion, at present for the sites investigated by radio echo sounding. The results suggest that a subglacial floodplain or valley delta has survived under a cold-based local ice divide in an ice sheet marginal zone for a long period of time, although the morphological interpretation of the bed topography is not totally indisputable. Short periods of warm-based ice coverage, allowing highly restricted erosion, may have occurred at this site during the evolution of the ice sheet. Nevertheless, during the Quaternary, cold-based, non-erosive conditions most probably dominated at the investigated locality, preserving the subglacial sediment accumulation. The glacially eroded valley reported in this paper, together with previous observations of subglacial cirques and U-shaped valleys, give further support for landform preservation beneath the East Antarctic Ice Sheet. Examples of calculated basal ice temperatures for sites on different altitudes within the same subglacial mountain massif demonstrate that localities at higher elevations have lower basal temperatures, which increases the probability for the occurrence of preserved morphology. The results of the present study indicate that no prolonged time periods of basal melting and glacial erosion have occurred at the Amundsenisen sediment locality during the ice sheet history.

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